Accumulation variability over a small area in east Dronning Maud Land, Antarctica, as determined from shallow firn cores and snow pits: some implications for ice-core records

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ABSTRACT. We investigate and quantify the variability of snow accumulation rate around a medium-depth firn core (160 m) drilled in east Dronning Maud Land, Antarctica (75°00′ S, 15°00′ E; 3470 m h.a.e. (ellipsoidal height)). We present accumulation data from five snow pits and five shallow (20 m) firn cores distributed within a 3.5–7 km distance, retrieved during the 2000/01 Nordic EPICA (European Project for Ice Coring in Antarctica) traverse. Snow accumulation rates estimated for shorter periods show higher spatial variance than for longer periods. Accumulation variability as recorded from the firn cores and snow pits cannot explain all the variation in the ion and isotope time series; other depositional and post-depositional processes need to be accounted for. Through simple statistical analysis we show that there are differences in sensitivity to these processes between the analyzed species. Oxygen isotopes and sulphate are more conservative in their post-depositional behaviour than the more volatile acids, such as nitrate and to some degree chloride and methanesulphonic acid. We discuss the possible causes for the accumulation variability and the implications for the interpretation of ice-core records.

1. INTRODUCTION

The spatial and temporal variability of accumulation can be caused by variations in ice-sheet climate, by the effects of snowdrift and sastrugi, by the movement of the ice sheet through a surface mass-balance pattern and by short-term climatic variations (Whillans, 1978). Several studies have been performed to estimate the snow accumulation in Dronning Maud Land (DML) within the European Project for Ice Coring in Antarctica (EPICA) DML pre-site survey programme. The accumulation rates on the high polar plateau range from ~45 to 90 mm w.e. a^{-1} (Holmlund and others, 2000), with low intercorrelation between different sites at short time-scales (Oerter and others, 1999, 2000; Karlöf and others, 2000; Sommer and others, 2000). The traverses have covered large areas where shallow and intermediate-depth cores have been drilled and snow-pit studies have been performed at different sites. Many of the sites have been connected with ground-penetrating radar, with the aim of following isochrones between the drill sites (Richardson and Holmlund, 1999; Eisen and others, 2002). Because of the large areal extent of the region under exploration, previous studies have concentrated on investigating accumulation variation on large scales (typically >100 km), and none of them has concentrated on the small-scale variability of the accumulation rate (e.g. 1–10 km scale). Given the large-scale influence of cyclones (King and Turner, 1997), a particular precipitation event by itself is unlikely to cause major variability of accumulation on km length scales. Most likely, the main cause of such variability is redistribution by snowdrift. Drifting snow is important for the local surface mass balance and may even lead to the formation of blue-ice areas, which can be regarded as islands of negative surface mass balance. For the mass balance of the whole Antarctic continent, snowdrift is only of importance in the coastal areas where snow is lost to the ocean (Loewe, 1970). For regional mass-balance measurements, the redistribution of snow influences the accumulation rate as measured from firn cores. Earlier work conducted during the 1996/97 Nordic traverse (Fig. 1) included firn coring, snow-pit studies, snow radar and meteorology (Winther and others, 1997). The most relevant

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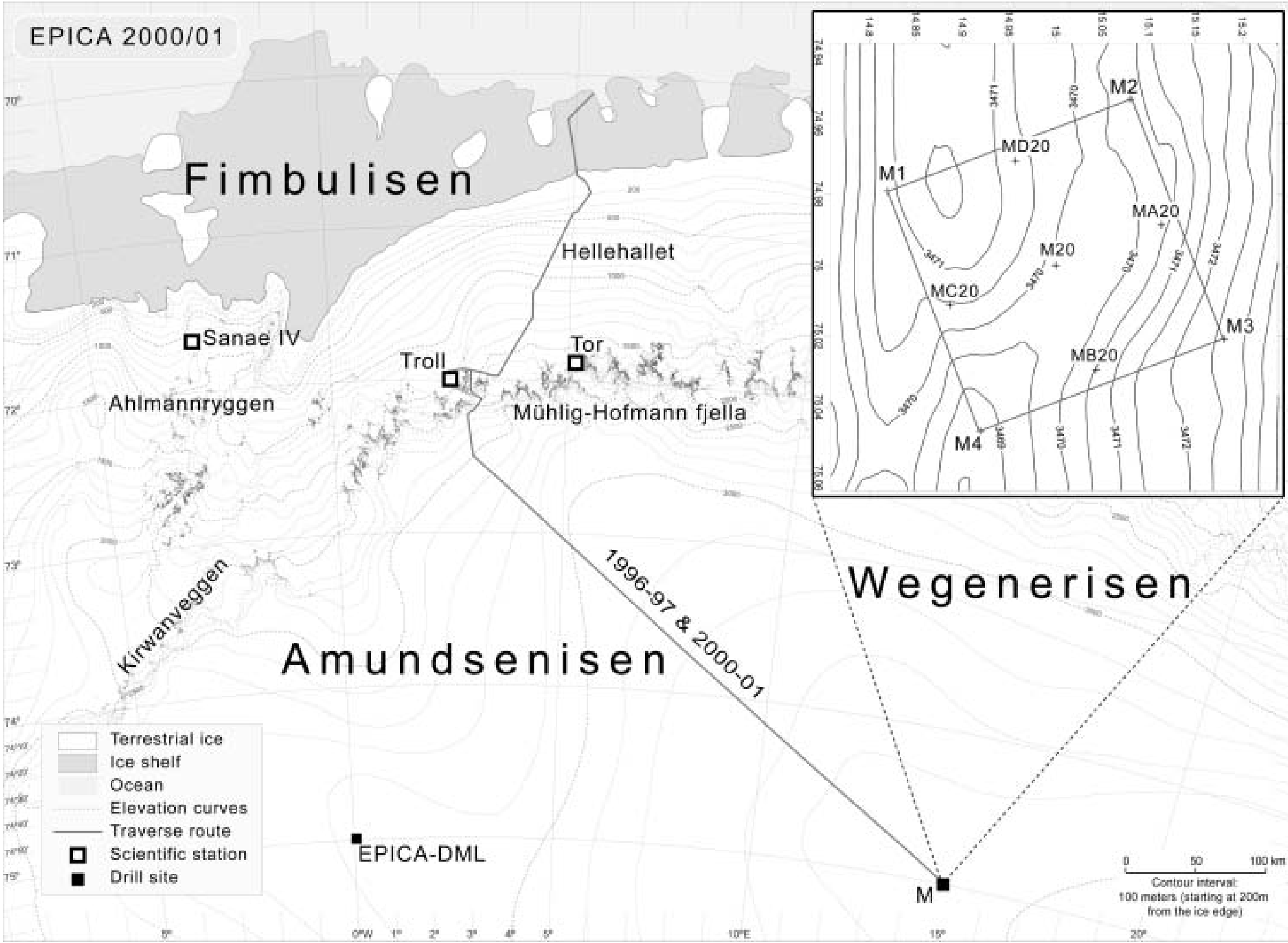

Fig. 1. The study area in DML, with the traverse route used in 1996/97 and 2000/01. The insert map shows drill locations (M20, MA20, MB20, MC20 and MD20) and the surface topography as measured by kinematic global positioning system (GPS) in the box limited by points M1–M4. The medium-depth core mentioned in the text was drilled ~5 m from M20. The EPICA-DML core site where a deep core is currently being drilled is also marked.

results of that work, for the purposes of this study, came from the 20 m core drilled at site M (Fig. 1) and the additional snow-pit sampling (Van den Broeke and others, 1999; Stenberg, 2000). The first study reported a mean annual temperature of –51.3°C and a mean accumulation rate for 1955–97 and 1965–97 of 51 ± 4 and 45 ± 4 mm w.e. a^{-1}, respectively.

In this work, we focus on accumulation rates derived from firn cores and snow pits retrieved in the 2000/01 summer season around a medium-depth drill site (site M) situated on the Antarctic Plateau (75°00′ S, 15°00′ E; 3470 m h.a.e. (ellipsoidal height)) (Winther and others, 2002). The main question addressed in this paper is whether we can detect accumulation variability from closely located snow pits and shallow firn cores. We further investigate how these variations can affect other firn-core and snow-pit time series and discuss the causes for the variations.

2. METHODS

2.1. Field methods

We sampled snow pits and shallow firn cores at five locations. The sites are named M, MA, MB, MC and MD, with site M in the centre (~5 m from the medium-depth core). The cores are named M20, MA20, etc. The distance from the centre to the auxiliary sites is 3.5 km (Table 1; Fig. 1). Each snow pit was sampled for density, ion chemistry, oxygen and hydrogen isotopes and stratigraphy to 1.5 m depth. This was deep enough to reach the 1991 layer, in which the Pinatubo (Philippines) and Cerro Hudson (Chile) eruptions left their tracers. The density was sampled with a vertical sample interval of 3 cm. To avoid the risk of disturbing the underlying snow during sampling, the snow was collected in a crossover pattern. Depth control was obtained by constantly levelling the sample depth with two adjacent rulers, to reduce the depth error. The estimated sample volume error was 1%, and the error in mass determination was $\pm 10^{-3}$ kg, yielding an uncertainty of 1.4% for the density. Samples for oxygen and hydrogen isotopes and ion chemistry were obtained at 2 cm intervals in the same pits. The number of samples per year was 5–15, which is sufficient to capture any annual signal. The pit wall was cleaned and marked at 2 cm intervals prior to sampling. We collected the samples from the bottom of the pit upwards, to prevent snow from higher up contaminating lower sampling levels. For every fifth sample, the depth was remeasured, and the anticipated depth was generally correct to within the same error as for the density samples.

The firn cores were retrieved with a Polar Ice Coring Office (PICO) lightweight coring auger in the snow pits, starting from a shallower depth than the bottom of the snow pit to allow for overlap between the pit and core records. When retrieving a firn core, there is always some unrecoverable loss of core material, predominantly in the upper metres. We used a correction scheme similar to the procedure described by Whillans and Bolzan (1988). Between 0.8 and 1.8 m of the cores could not be used due to low core quality. The cores were drilled to a depth covering the period back to the eruption of Tambora, Indonesia, in AD 1815 and the unknown eruption in AD 1809. This well-known double peak (Cole-Dai and others, 1997) has served as the main time horizon for the dating of the firn cores. This peak was found around 7.6 m w.e. in the cores.

The solid conductivity of the firn cores has been measured with electrical conductivity measurement (ECM; Hammer, 1980) and dielectric-profiling (DEP; Moore and Paren, 1987; Moore and others, 1989; Moore, 1993; Wilhelms and others, 1998) instruments, and their concentration of radionuclei (^{137}Cs) and oxygen isotopes (^{18}O/^{16}O) has been measured. We used ~300 mm and 50 mm sampling intervals for the radionuclei and oxygen isotopes, respectively. For the time series, the results are linearly interpolated for the areas where no core data are available.

The yearly accumulation rate (b) is derived from the ratio of water equivalent (w.e.) depth or mass of firn column above the dating horizon to the time-span covered by the firn column. The associated relative errors can be expressed as

$$\frac{\delta b}{b} \leq \sqrt{2 \times \left(\frac{< \frac{1}{2}\mathrm{sw}}{\Delta \mathrm{depth}}\right)^2 + \left(\frac{\delta y_{\mathrm{diff}}}{y_{\mathrm{diff}}}\right)^2 + \left(\frac{\delta D_{\mathrm{diff}}}{D_{\mathrm{diff}}}\right)^2}, \quad (1)$$

where sw is sample length, y_{diff} is time difference for the dating horizons used (e.g. uncertainty in the date of volcanic deposition) and D_{diff} is depth differences in m w.e. between dating horizons.

When the sample length is large compared to the depth between dating horizons, the first term on the righthand side is important. For high-resolution measurements, however, such as electrical measurements, the first term can be neglected. The typical error in b is ~9% and ~6% for the snow pits and firn cores, respectively.

The error estimate only applies at the identified dating horizons. Any time-scale (i.e. accumulation) variation between the dating horizons is not captured by this error estimate.

2.2. Laboratory methods

For the δ^{18}O measurements of snow-pit samples, performed at Centre for Isotope Research (CIO) Groningen, the standard deviation of the measurements was ±0.10‰ (^{18}O‰). The overall accuracy is only slightly higher. This estimate is based on working standards, and on the results for samples analyzed in duplicate.

δ^2H measurements have also been conducted, but for the present work, δ^2H is used only to support the δ^{18}O data, and to discriminate between 'real' and sampling/analysis-based outliers in the records.

The oxygen isotope samples from the firn cores were analyzed by the stable-isotope mass spectrometer at the University of Copenhagen. This is a similar set-up to that used at CIO, although the reproducibility is slightly better (±0.05‰). All δ^{18}O values reported here are expressed on the Vienna Standard Mean Ocean Water–Standard Light Antarctic Precipitation (VSMOW–SLAP) scale (Gonfiantini, 1984), i.e. calibrated using VSMOW and normalized using SLAP. The latter step is especially important here, since the isotopic composition of the firn is similar to that of SLAP.

Table 1. Positions and ellipsoidal elevation of drill sites and the adjacent snow pits. The elevation is determined by kinematic GPS measurements with a local base station situated at site M. The drill depth for each core is also given

Site	Lat. (S)	Long. (E)	Elevation m h.a.e.	Drill depth m
M	74°59′59.9″	14°59′46.9″	3470.0	19.22
MA	74°59′19.4″	15°06′48.2″	3470.4	20.63
MB	75°01′45.7″	15°02′36.6″	3470.5	20.12
MC	75°00′40.3″	14°53′11.3″	3470.4	20.17
MD	74°58′14.3″	14°57′24.0″	3470.8	21.27
M 96/97	74°59′59″	15°00′06′	3470.0	

The British Antarctic Survey has performed measurements of the following ions: sodium (Na^+), potassium (K^+), magnesium (Mg^{2+}), calcium (Ca^{2+}), fluoride (F^-), methanesulphonic acid (MSA) ($CH_3SO_3^-$), chloride (Cl^-), nitrate (NO_3^-) and sulphate (SO_4^{2-}). The instrumental set-up is described in Littot and others (2002). The reproducibility of the measurements is 4–10% depending on the species and concentrations (Littot and others, 2002).

The electrical measurements on the firn cores were performed in a field laboratory. The laboratory was excavated from the snow, and the roof was made of wood covered with snow to increase the albedo. This was sufficient to maintain a stable temperature (–23°C) in the laboratory during working hours. The resolution of the electrical measurements was 2 mm for ECM and 5 mm for DEP. The DEP measuring electrode was 10 mm and the measurements were performed at 250 kHz. Both ECM and DEP measurements have been temperature-corrected to –15°C.

^{137}Cs-radioactivity measurements were performed by the Laboratoire de Glaciologie et Géophysique de l'Environnement (LGGE) to determine the accumulation for the last 35–45 years using the well-known radioactive reference layers of 1965 and 1955 (Delmas and Pourchet, 1977; Pinglot and Pourchet, 1979).

2.3. Dating of snow pits and firn cores

The density, oxygen isotope and ion chemistry data were used to date the snow pits. In the non-sea-salt (nss) sulphate record, the reference horizons are provided by the deposition from the Pinatubo and Cerro Hudson eruptions. The aerosol clouds from Cerro Hudson were centred over the South Pole in September 1991, and indications of the arrival of the Pinatubo plume were noted in November 1991 (Cacciani and others, 1993; Saxena and others, 1995). By using mainly seasonal cycles in the oxygen isotope data, supported by comparison with ion records, an attempt was made to construct a yearly accumulation record for each snow pit over the last 9 years (Fig. 2).

The dating of the firn cores followed the same analysis scheme as outlined in Karlöf and others (2000), using the electrical records to find volcanic dating horizons. Both the DEP and ECM records were de-trended by removing the mean, then low-pass filtered. We then created a high-pass version by subtracting the low-pass filtered data from the

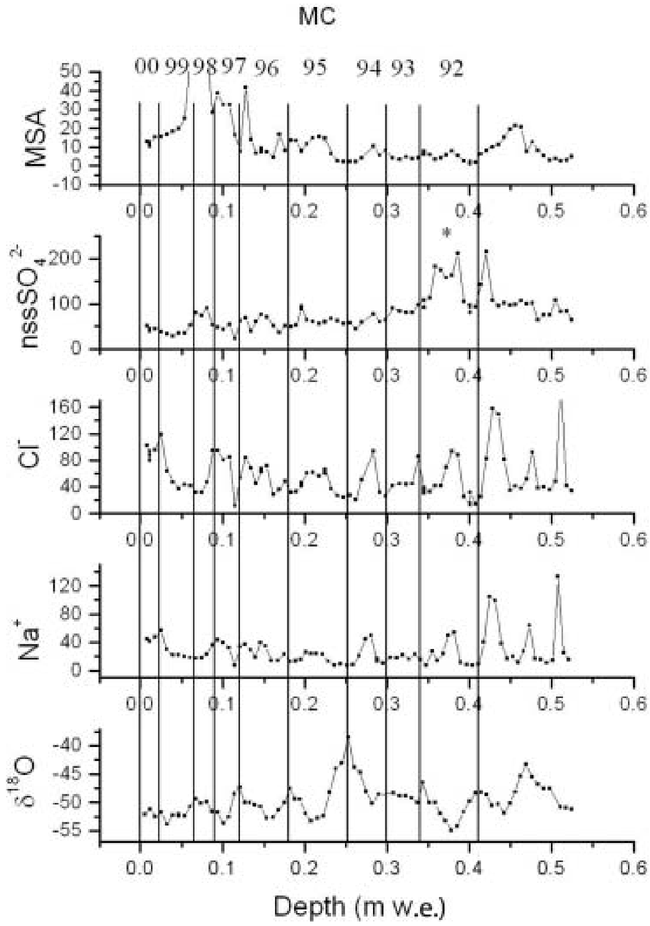

Fig. 2. Example, from pit MC, of how several different species have been used to date the pits. We have mainly used oxygen isotope data, with the support of ions. Year transitions are marked with vertical lines, and the 1991 Pinatubo lower key horizon is marked with a *.

Table 2. Average accumulation values (10^{-3} m w.e. a^{-1}) obtained from the snow pits. The error in estimated snow accumulation for each pit is derived from error propagation. The error stated for the area mean is the standard error of the mean

	Snow pit					
	M	MA	MB	MC	MD	Area mean
Mean	50 ± 1.1	48 ± 1.3	46 ± 1.0	46 ± 1.1	53 ± 1.2	49 ± 1.3

original. Thus, we have removed unknown low-frequency relationships, such as the density effect on the ECM signal in the measurements. Finally, the residual was low-pass filtered at a higher frequency (called high-pass filter in Karlöf and others (2000)) and the output was used for further analysis. Both the ECM and DEP output were then normalized to one standard deviation, and peaks having positive amplitude over 2σ were considered for dating purposes (Fig. 3). The variance of the dataset was estimated using all the sample data; the peaks were not omitted. This leads to an increase in σ and therefore a higher and more conservative threshold value. The well-known eruptions of Tambora (AD 1815), identified in the cores, and Pinatubo (AD 1991), identified in the snow pits, were used as key horizons and helped to steer our search for other possible volcanic dating horizons. The number of peaks identified in each core varied from 20 (M20) to 11 (MC20), but six peaks were identified in all cores. These six horizons are correlated to well-known volcanic eruptions and were used to calculate the accumulation rate between these eruptions. The uncertainty in the dating is due to the uncertainty of the age of the volcanic horizons used (<1 year).

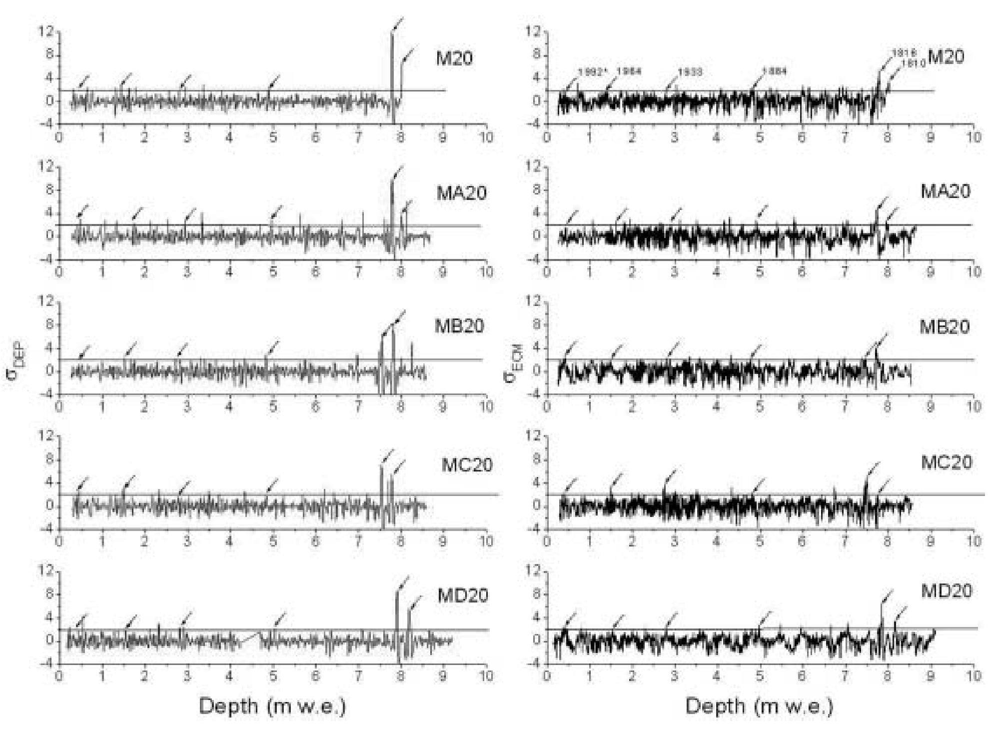

Fig. 3. ECM (right) and DEP (left) data from the firn cores. The data are normalized to one standard deviation. Peaks exceeding a 2σ threshold are considered to be of volcanic origin. Six volcanic peaks were identified in each core and used for dating. These peaks are listed in Table 5. Depth of peak is derived from the snow pits.

Table 3. Mean, standard deviation (SD) and standard error (SE) of the oxygen time series covering the full length of the firn cores and snow pits.

Core	Mean	SD	SE	Number of samples
	‰	‰	‰	
M20	–49.94	±1.79	±0.098	334
MA20	–50.01	±1.87	±0.098	357
MB20	–49.88	±1.88	±0.099	360
MC20	–50.04	±2.08	±0.111	349
MD20	–49.81	±1.86	±0.093	399
M 96–97*	–49.49	±1.97	±0.097	406
Snow pit				
M	–49.94	±2.49	±0.298	70
MA	–50.05	±3.14	±0.362	75
MB	–49.50	±2.67	±0.312	73
MC	–49.77	±2.94	±0.342	74
MD	–49.13	±3.34	±0.386	75

*Firn core drilled in the same area during the 1996/97 field season (Isaksson and others, 1999).

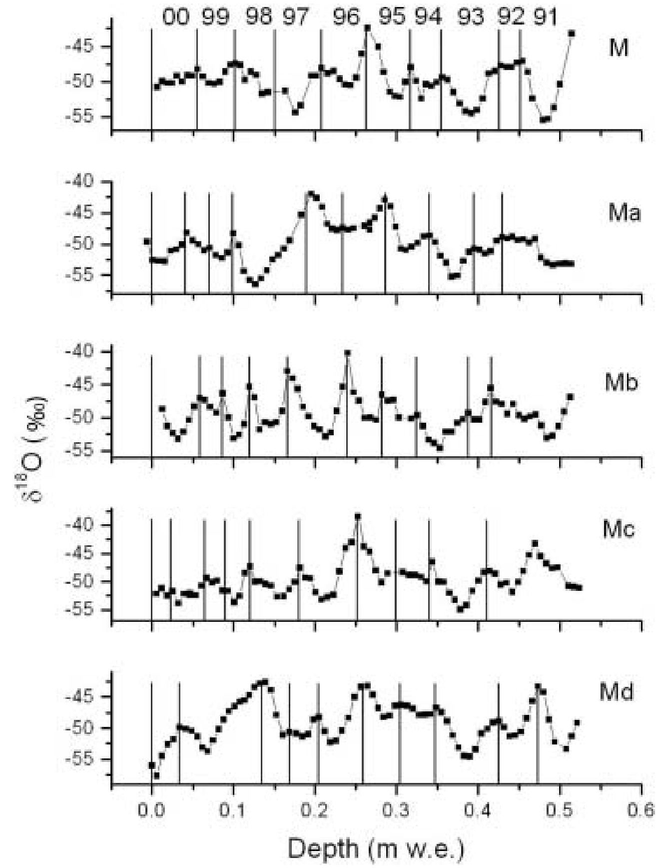

Fig. 4. Oxygen isotope data for the five pits. The year transitions are marked with vertical lines.

3. RESULTS AND DISCUSSION

Data obtained from snow pits and firn cores are affected by several factors. The source for the data can be represented by a signal with an underlying dynamic and noise. Wet and dry deposition events at the snow surface can be regarded as sampling from this signal. Post-depositional processes at the surface and at depth finally determine the data quality of the recorded species. The sampling resolution of the snow in the pit and along the core determines the time resolution.

3.1. Snow pits: the last 9 years

The sampling date and the nss-sulphate peak (Pinatubo, AD 1992) were used as key horizons for determining the average accumulation rate over the last 9 years (Fig. 2). An attempt was also made to date the pits on a yearly basis. For this exercise the most positive value in each 'year' cycle in the oxygen isotope records was used to indicate a year transition. The ion records were used as an independent check: sodium and chloride, which both peak in late winter (Bergin and others, 1998), are anticorrelated with the oxygen isotopes for the years with clear cycles (Fig. 2). Physical stratigraphy was not used as a dating parameter, due to the absence of clear depth-hoar layers. When there was a clear dampening in the oxygen isotope signal, which was often reflected in the other records, a year transition was chosen that would not create more variance between years. Due to these shortcomings, a full yearly accumulation rate record is not presented. However, where there are clear cycles in the oxygen isotope records (e.g. 1997 in pit MA and 1999 in pit MD) it is evident that there is both year-to-year variability (temporal) within the records and variability (spatial) between the sites (Fig. 4). Accumulation values derived from snow height measurements at a nearby automatic weather station (AWS; ~100 m from site M) support this interpretation (Reijmer, 2001). The fact that several records obtained from the same pit show dampening in the same depth interval indicates that there is an external factor, perhaps the redistribution of snow, determining the shape of the record (Fig. 2, years 1993, 1996 and 1999).

The 9 year mean accumulation rate is summarized in Table 2. The pit with the lowest accumulation rate is situated on the 'steepest' slope in the experiment area. This observation is consistent with results from snow radar measurements performed in the area (Karlöf, 2004). However, none of the accumulation rate estimates for the last 9 years is significantly different statistically from the others.

3.1.1. The oxygen isotope records

The stratigraphies as depicted by the oxygen (and hydrogen (not shown)) isotope records (Fig. 4) show variations in amplitude and period with time. For example, MB shows annual cycles with less variation between years, for the upper part of the pit (<0.2 m w.e.), and MC shows a clear amplitude increase at 0.25 m w.e., but for the rest of the record the interannual amplitude variance is smaller (Fig. 4). The mean and variance for each pit are listed in Table 3. The isotopic values of meteoric precipitation are correlated with temperature, and they also vary with altitude, latitude and proximity to the ocean (Dansgaard, 1964). In seeking to explain the differences between the time series, we can neglect all these factors since the sites are all located at practically the same point. Thus, variability between the recorded series must be attributed to local processes such as wind redistribution of snow and pumping of air within the snowpack. However, from the mean (~10 years) isotopic value for the different pits, we can conclude that the mean differences are not significant for any of the pit combinations.

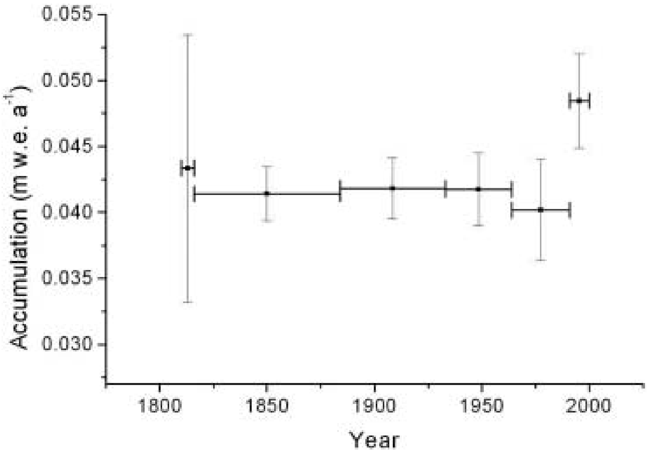

Fig. 5. Mean accumulation record for the five cores and their 95% confidence interval. The data based on radioactive measurements are excluded.

3.1.2. The ionic record

All the ion records show different characteristics in terms of mean and variance, both over the whole period and for parts of it (Fig 2; Table 4). The discrepancy between the various ion records can be attributed to the variability created by the changing snow surface, by depositional noise (Fisher and others, 1985) and post-depositional processes. The noise can be assumed to be on the same spatial scale as the main surface patterns in the area (sastrugi ~2–10 m and ice-sheet undulation ~1–2 km). To reduce the impact of depositional and surface noise in our records, we compare the mean over 10 years (Table 4). The comparison is made by testing for a statistically significant difference between the 10 year averages of the concentration in the pits. When few pit combinations show significant difference in mean, the species has a homogeneous spatial distribution (i.e. conservative behaviour). For the sea-salt ions (chloride, sodium and potassium), pit M stands out as being significantly different (higher concentrations) from the other pits. However, the difference between the means of the MC and MD chloride records is significant at the 95% confidence level. Sulphate is the one ion with only one combination (MD–MC) that shows a significant difference between the means, i.e. it shows a more conservative behaviour than the sea-salt ions.

The MSA mean for MD is significantly different from that of the other pits except for MA. This is due to the different pattern in the MSA signal at shallow depth. Nitrate, known to be affected by re-emission to the atmosphere, shows significant differences for most of the combinations, and the pairs that are not rejected are close to the threshold, except for the M–MD combination. Nitrate shows high uniformity over large parts of the continent when only snow below 1 m depth is considered (Mulvaney and Wolff, 1994). Our results are greatly affected by the low accumulation rate and thus potential re-emission of nitrate to the atmosphere. The mean ion concentrations from a snow pit sampled from the same area during the 1996/97 field campaign reveal values in the same range as for the present results (Stenberg, 2000).

Whatever the mechanisms involved in the deposition of the species – wet or dry deposition, or post-depositional processes and their mutual apportionment – we can conclude from the analysis that some species have higher spatial variance than others. It also seems that the species affected by higher spatial variance (less conservative) are not irreversibly trapped in the snow like MSA and nitrate.

3.2. Firn cores: the last ~200 years

Accumulation time series derived from shallow firn cores are shown in Figure 5. The accumulation rate values are averages between the identified dating horizons (Table 5). No accumulation values are distinctly different from others, indicating that the values obtained for longer periods are more robust with respect to spatial variability than those derived from the pits. Figure 5, depicting the area mean accumulation rate for different periods and the associated standard error of the mean, indicates that the two periods covering the 19th and early 20th centuries (1816–84 and 1884–1933) show stable values, with similar spatial variance between the periods (i.e. similar error-bar width). The period 1933–64 shows slightly higher variance but no change in accumulation, the period 1964–92 suggests a lowering, although not significant, in accumulation, and the period covered by the pits shows up to ~17% higher accumulation than the previous period. Accumulation values based on radioactive measurements were calculated for 1955/65–2000 and 1955–65, respectively. These results show a decrease from the mid-1960s onward (not shown). Overall, these results show that the accumulation rate in the area has been stable over the last 200 years.

Table 4. Statistics of the ion records from the snow pits (concentrations are in ppb). Italic numbers are based on values close to our detection limit

Ion	Snow pit										Area mean	
	M		MA		MB		MC		MD			
	Mean	SD	Mean	SD	Mean	SD	Mean	SD	Mean	SD	Mean	SE
MSA^-	12.9	13.3	10.8	19.7	13.9	14.4	13.5	15.7	9.0	6.8	12.0	0.9
Cl^-	60.9	39.1	48.5	31.0	49.5	23.5	56.1	32.5	44.3	22.0	51.8	2.9
NO_3^-	65.0	55.3	55.4	23.9	51.5	23.4	41.9	16.3	69.5	36.0	56.7	4.9
SO_4^{2-}	97.6	41.0	85.9	54.8	90.9	53.2	85.3	38.1	99.9	51.5	91.9	2.9
$nssSO_4^{2-}$	88.3	41.7	80.9	54.4	85.8	52.8	79.6	37.7	95.3	50.7	86.0	2.8
Na^+	36.1	30.8	21.7	14.6	22.0	14.3	27.6	22.0	21.7	17.1	25.8	2.8
K^+	10.4	18.7	3.5	5.6	3.7	1.8	4.4	5.6	5.4	6.7	5.5	1.3
Mg^{2+}	*1.9*	*2.3*	*2.0*	*2.3*	*2.4*	*2.2*	*3.4*	*3.4*	*1.6*	*2.6*	*2.2*	*0.3*
Ca^{2+}	*4.7*	*4.8*	*4.1*	*4.4*	*5.1*	*3.8*	*6.2*	*7.6*	*7.2*	*5.9*	*5.5*	*0.6*

Although the accumulation rates do not cover exactly the same time period, they are consistent with the values recorded in a previous study (Van den Broeke and others, 1999) (Table 5).

3.2.1. The oxygen isotope record

The mean and variance of the full oxygen isotope record (~200 years) for the five firn cores are shown in Table 3. The mean values over the entire core length for all cores are slightly more negative than the value reported for the same area in an earlier study (Isaksson and others, 1999). The cores in both studies cover approximately the same period except for the most recent years. The mean isotope values of the pits cannot explain this difference. Three of the pits record more positive (higher) values than the cores, and two of the pits roughly the same mean value as the corresponding core. Higher values in recent years should shift the mean of the cores towards less negative values.

The variability in the ~180 year mean isotope value between the cores probably reflects how small differences in depositional and post-depositional environments affect the oxygen isotope records. As for the snow pits, the mean and variance vary with time at the different sites, and localized (in time) events in the different records do not show up in all cores (Fig. 6). However, the pronounced event around 1950 at site M is visible in all the cores except MD20, although highly attenuated in the other cores compared to M. The record from core MD20 shows only a small step around 1950 in the slope in the corresponding time period, which might stem from the same event but very attenuated. Other examples of the differences in mean and variance of the firn records are two marked events in MA around 1850 and 1810 which are absent from some of the other cores and, where present, are not as pronounced. We interpret these differences as impacts of the local environment (see section 4) and they are thus an indication of how such processes affect oxygen isotope records over a period of 5–10 years (Fig. 6). Despite the discrepancies noted, the means of the full oxygen isotope records (~200 years) do not vary significantly.

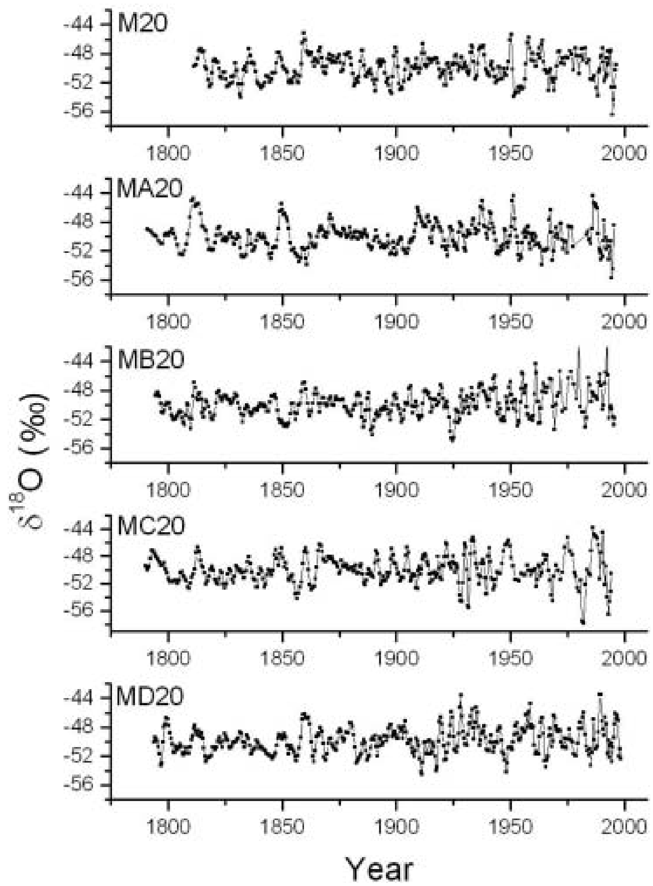

Fig. 6. Oxygen isotope records from the five cores covering ~200 years. Note the difference between the series around 1810 and 1850 (see text for details).

4. CAUSES OF ACCUMULATION VARIABILITY AND IMPLICATIONS FOR FIRN-CORE TIME SERIES

Accumulation variability is caused by processes that act on different spatial and temporal scales. Processes relevant for this study are the effects of sastrugi, drift and short-term climatic variations. Topography dominates the wind

Table 5. Accumulation values (10^{-3} m w.e. a^{-1}) based on the firn cores and overall mean for the area. For all 'Area mean' values the standard error of the mean is given. The period 1992–2000 is derived from the snow pits. Each row represents the mean accumulation rate for the given period. The lower part of the table represents the accumulation values derived from radioactive measurements. The values from Van den Broeke and others (1999) are added for comparison

		Firn core					
Period	Volcanic eruption	M20	MA20	MB20	MC20	MD20	Area mean
1992–2000	Pinatubo, Philippines	50 ± 1.1	48 ± 1.3	46 ± 1.0	46 ± 1.1	53 ± 1.2	49 ± 0.8
1964–92	Agung, Indonesia	36 ± 2.5	44 ± 3.0	38 ± 2.7	41 ± 2.8	41 ± 2.8	40 ± 1.4
1933–64	Cerro Azul, Chile	45 ± 3.0	42 ± 2.8	43 ± 2.9	40 ± 2.7	39 ± 2.7	42 ± 1.0
1884–1933	Krakatau, Indonesia	42 ± 2.8	41 ± 2.7	40 ± 2.7	42 ± 2.8	45 ± 3.0	42 ± 0.8
1816–84	Tambora, Indonesia	43 ± 2.9	42 ± 2.8	39 ± 2.6	40 ± 2.7	42 ± 2.8	41 ± 0.7
1810–16	unknown	40 ± 4.2	33 ± 4.0	50 ± 4.7	40 ± 4.2	53 ± 4.8	43 ± 3.7
Accumulation rates derived from radioactive measurements							
1955–2000		40 ± 4.5	42 ± 4.7	42 ± 4.6	39 ± 4.5	40 ± 4.5	41 ± 0.5
1965–2000		30 ± 5.5	42 ± 5.8	37 ± 5.3	42 ± 6.0	45 ± 6.2	39 ± 2.5
1955–96*		51 ± 4					
1965–96*		45 ± 4					

*Van den Broeke and others (1999).

redistribution of snow such that the general snow distribution pattern is more sensitive to changes in wind direction than in wind speed (Liston and Sturm, 1998). However, on the Antarctic Plateau the large-scale surface undulations are of such low frequency (long wavelength) that they influence the wind pattern, the katabatic flow. On the other hand, the high-frequency (short-wavelength) topography, consisting of small ridges, dunes and sastrugi, changes with the redistribution of the surface snow. Thus, the important small-scale topography is continually varying, leading to variable conditions that influence the small-scale pattern of snow distribution and accumulation. The question is whether persistent katabatic flow or varying wind direction due to *synoptic* activity is more important. In winter, when the surface cooling is strong, gravity-driven winds dominate, whereas in summer, when the surface cooling is weaker, the large-scale pressure-gradient force dominates (Kodama and others, 1989; Van den Broeke and others, 2002). Model simulations with a high-resolution global climate model, studying persistent flow features, suggest that the annual mean directional constancy in the area is 0.5–0.75 (Van den Broeke and others, 1997). Since the 1996/97 traverse, one AWS has been positioned at site M, and the data suggest even higher directional constancy, 0.8 (Reijmer, 2001). The direction of the dunes and sastrugi follows the main easterly wind direction. However, there is a crossing sastrugi/dune pattern which might be an indication that the katabatic flow is influenced by the gradient wind determined by the synoptic activity. Winds with high directional constancy remove the snow from the area, or, if convergence occurs in the area, snow from higher altitudes accumulates. Low directional constancy presumably leads to mixing of the original snow and low influx of material from other areas.

For snowdrift to occur, the lifting capacity in the wind has to be larger than the snow-holding capacity (the surface shear strength). The surface shear strength depends on the stage of metamorphism of the surface snow. The surface pattern and variations in snow surface density (Winther and others, 2002) cause the snow to be eroded in a non-uniform way, leaving stripes of snow and bands of eroded areas. The scale of this can vary from a few metres to kilometres. An illustrative 'extreme' example of this feature is shown in Liston and others (2000, fig. 2). The authors of the current study have observed features of this kind (10–100 m scale) both at the study site and at site 'Camp Victoria' (76°00′ S, 08°03′ W) further west. If no more erosion occurs until the next precipitation event, this pattern is buried and preserved in the snowpack. How this shows up in a snow pit is depicted by Stenberg and others (1999). The intersections of layers indicate the boundary of such a snow stripe. Records obtained from an area where this kind of process occurs will show differences; for example, records of 'pulsed' events like volcanic aerosol in ECM and DEP will presumably show different amplitudes of the pulse depending on whether the site has been eroded. This explains the differences in peak amplitudes between the electrical records (Fig. 3). The kind of erosion described above causes an increase in spatial variance for all species that have been trapped in the snow. If one assumes that the process of dune and sastrugi formation is random within certain limits, an average value based on several measurements retrieved from a small area will improve the estimate of the sampled species.

The negative mass-balance term, sublimation, is unlikely to cause significant net accumulation variability as long as the wind and atmospheric moisture field over the domain is uniform (Van den Broeke and Bintanja, 1995). However, sublimation is important for the re-emission of chemical species to the atmosphere, especially at sites like this with low accumulation, and thus has an important effect on the ionic record (De Angelis and Legrand, 1995; Wagnon and others, 1999).

As shown earlier, the accumulation variability in the experiment area is only significant for a few periods. This suggests that time series of wet deposited species (e.g. isotopes and sulphate) should show a high correlation. This was shown in section 3 by comparing the differences in 10 year means of the analyzed species. The oxygen isotopes and sulphate show no significant differences, but other of the studied species seem to be affected not only by accumulation variability. We conclude that post-depositional processes are of major importance for the variability of the analyzed species. The spatial scale of the variability of post-depositional processes depends on the physical processes involved (e.g. exposure of the surface due to erosion of snow facilitates re-emission of ions). It is also determined by the local sastrugi/snow-dune pattern, which also determines the effect of snow ventilation (Waddington and others, 1996). Snow ventilation is an important mechanism for post-depositional alteration of the oxygen isotope signal (Neumann and Waddington, 2004) and is a possible cause for the differences in oxygen isotope records on short time-scales.

5. CONCLUSIONS

The multi-parameter analysis performed in this study allows us to quantify accumulation variability between core and pit sites located close to each other (i.e. on the km scale). In addition, we suggest possible causes for the variation and we have studied how these variations affect commonly collected time series. For some years, distinct year-to-year accumulation variability between the snow pits was detected, but there was no significant difference when the 9 year mean was considered. The lowest accumulation rates were found at sites located at the 'steepest' slope. No periods of distinct changes in accumulation are found in the firn cores, mainly because of the long averaging periods. The stacked accumulation record shows no trend in accumulation over a ~200 year period. Stacking the estimates for the sites allows us to calculate confidence intervals to facilitate the interpretation of the spatio-temporal accumulation variability at our study site. The spatio-temporal variability is naturally low for the longer averaging periods, but on a year-to-year scale it can increase substantially. We suggest that the recorded variability in accumulation is part of the natural noise contained in the accumulation process. Thus, by averaging the accumulation rates spatially and in time, the values obtained will be more robust. Since the records are collected close to each other, it is assumed that a climatic signal is still present in the record despite the averaging.

When investigating how the accumulation variability affects ion chemistry and oxygen isotope (and hydrogen) time series, we find that oxygen isotopes and sulphate show less spatial variability than nitrate and, to some degree, MSA and chloride. Specific events in the ion chemistry and isotope time series do not correlate with the other sites in any of the time series. We therefore believe that locally steered post-depositional processes are of major importance for the short-time-scale, spatial variability of the analyzed species. This has implications for detailed analysis of firn-core time series obtained from sites with low snow accumulation rate.

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